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T R A N S L A T I O N

Number 66 28 Apr 1947

ARTICLES ON METEOROLOGY AND HYDROLOGY

Prepared by

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S U M M A R Y O F C O N T E N T S

Articles on Meteorology and Hydrology
(Doc No 458226)

This is an extract translation composed of full translations of seven articles which appeared in the January 1946 issue of the periodical "Meteorology and Hydrology," published by the Central Administration of the Hydro-meteorological Service of the Council of Ministers, USSR.

The articles, contributed by Soviet scientists, discuss various meteorological and hydrological experiments and observations. They are amply illustrated with tables, graphs, diagrams and maps.

Pages 1 through 48

[redacted]

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MEMORANDUM FOR THE ASSISTANT CHIEF OF STAFF, G-2, GSUSA
Attn: Chief, Security Branch, Signal Corps

SUBJECT: Downgrading of Report.

REFERENCE: Your Memo SIGSS, dated 16 March 1951, same subject.

1. In response to reference the below listed document is declassified provided all CIA identification is removed.

CIG Documents Branch Translation, No. 66
(Articles on meteorology and hydrology)
dated 28 April 1947.

FOR THE ASSISTANT DIRECTOR, COLLECTION & DISSEMINATION

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ARTICLES ON METEOROLOGY AND HYDROLOGY

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January 1946

EDITOR'S NOTE: Bibliographies appear at the end of articles I, II, III and V. References in the text to the bibliographies are indicated by a number in parentheses.

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Approved For Release 1999/08/25 : CIA-RDP78-03109A000500010007-7

I. MEASUREMENTS OF EFFECTIVE RADIATION OF WATER VAPOR IN THE RANGE OF MAXIMUM TRANSPARENCY

by

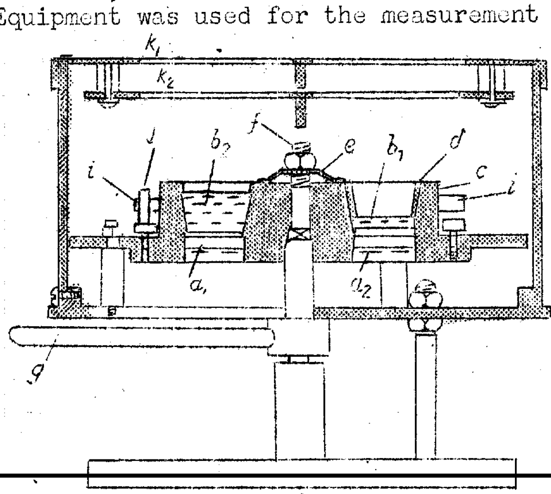
V. Kastrov

EDITOR'S NOTE: The following is in explanation of the Russian symbols as used in the text:

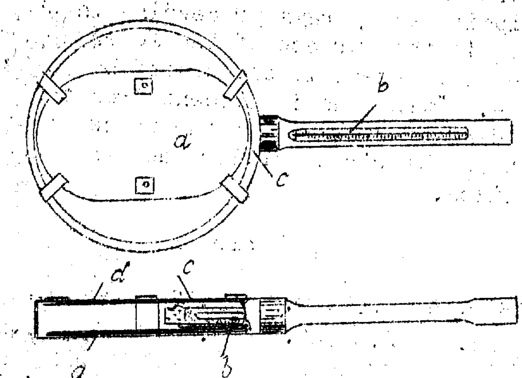
П as in f	=	dust, Russian П Н Л
кПВ	=	continental, polar, high.
ПрАВ	=	transitional, arctic, high.
кТВ	=	continental, tropical, high.

Effective radiation is usually measured for a freely exposed horizontal plane and in the whole spectrum. It is rather difficult, however, to interpret this quantity theoretically. From the theoretical point of view, measurements in a sufficiently small solid angle and for narrow bands of the spectrum would be of the greatest interest. We have certain empirical data for the distribution of diffusion in the sky. Such data however, are almost entirely lacking for spectral composition, except for the measurements of Fowel with filters of fluorite and rock salt, which were of tentative nature.

The range of the greatest interest, where water vapor has maximum transparency of basic absorbtion and radiation in the atmosphere, is approximately 8.5 - 12μ . Therefore, special measurements of effective radiation were made in the Kuybyshev Geophysical Observatory with a cloudless sky and in a band of the spectrum with limits approximating these wave lengths. They were made from June 1944 to January 1945. Isolation of the required spectral band was accomplished by two consecutive measurements: One time with a thinner, and the other time with a thicker fluorite light filter. Equipment was used for the measurement as in figure 1.



Here A_1 and A_2 are thermopiles of Yanishevsky's system with a small lag (about 20 seconds) and great sensitivity. With the object of making an instrument which will not be very sensitive to short wave radiation for measuring in the dark, they are colored with magnesium oxide, the coefficient of absorption of which is very high in the remote intra-red region and can be regarded as independent of λ. b_1 and b_2 are fluorite filters, 2 and 8 mm thick, in the form of discs with a slightly conical lateral surface. C is a solid brass plate serving as a setting for the filters. Their lateral surfaces are ground to fit its sockets. The filters are inserted in the sockets loosely and held in place by the rings of plate d and spring e. The whole setting can be turned on axis f to 180°, for which lever g is used, so that any kind of filter can be placed over each of the thermoplies. The accuracy of the adjustment of the setting is guaranteed by screw i of the setting and by pin j. K_1 and K_2 are diaphragms. The setting and the plate with rings are thoroughly chrome plated, so that their temperature will not vary much with the opening and closing of the top of the instrument.



The top consists of a brass plate a (figure 2), painted black, the temperature of which is determined by thermometer b. The reservoir of the thermometer enters into copper pipe c, soldered to plate a and packed with metal shavings. Plate d of polished aluminum, protects plate a from recooling due to radiation from the sky. The aperture of the instrument is limited by the openings of the diaphragm K_1. Therefore, not only the beams which proceed directly through the diaphragm and filter, but also those which are reflected from the inner surfaces of the sockets and rings of plate d (figure 1) fall on the thermopile. Since the theory and methods of measuring effective radiation with light filters are not yet worked out, we must pay particular attention to them in the present article.

The intensity of radiation absorbed by the thermopile---radiation per solid angle at angle z for spectral band $d\lambda$---may be expressed by the formula:

$$di_{\lambda,j} = b(\lambda,z)d\lambda aDt_{\lambda,j}(z)S\Phi_j(z)\cos z, \quad (1)$$

where $b(\lambda,z)$ is the monochromatic intensity at the incident angle; a is the coefficient of absorbtion of the thermopile; D is the factor of reflection of the filter; $t_{\lambda,j}(Z)$ is its coefficient of transmission; S is the area of the thermopile; and $\Phi_j(z)$ is the coefficient, taking account of additional light from its rays reflected from the surface of the ring in the socket. We shall designate the thin filter by the index $j=1$, and the thick one by $j=2$.

It is not difficult to show that if we overlook in the first approximation the very small loss of energy accompanying reflection from the chrome surface and consider that the reflected rays also fill the lower aperture of the socket of the filter, then,

$$\Phi_j(z) = \frac{S_{2j}}{S_{1j}}, \quad (2)$$

where S_{2j} is a part of the surface of the upper aperture of the ring which is illuminated by rays proceeding in the given direction, and S_{1j} is the surface of the lower aperture of the setting. The area S_{2j} is easily calculated by the corresponding geometric formulae. $\Phi(z)$, as is evident from table 1, does not have completely identical values for the thick and thin filters in our instrument as the rings holding them in place have been to be of different sizes.

Table 1

cos z	1	0.975	0.950	0.925	0.900	0.875	0.850	0.825	0.800	0.775	0.750
z°	0	12.8	18.1	22.3	25.8	29.0	31.8	34.4	36.9	39.2	41.4
$\Phi_1(z)$	1.03	1.03	1.03	0.99	0.86	0.68	0.53	0.37	0.21	0.09	0.02
$\Phi_2(z)$	1.13	1.13	1.13	1.07	0.94	0.74	0.56	0.40	0.24	0.12	0.03

The complete intensity of the radiation absorbed by the thermopile will be:

$$i_\lambda = \int_\lambda\int_\omega di_{\lambda,j}dw = 2\pi aDS\int_\lambda\int_z b(\lambda,z)t_{\lambda,j}(z)\Phi_j(z)\cos z \sin z dz\, d\lambda,$$

or, considering that $b(\lambda,z)t_{\lambda,j}(z)$ is not significantly changed within the aperture of the angle of the instrument,

$$i_j = \pi aDS\Psi_j\int b(\lambda,\bar{z}_j)t_{\lambda,j}(\bar{z}_j)d\lambda,$$

where $\bar{z}_j$ is a certain median value of z (center of gravity of the aperture), and $\Psi_j = 2\int \Phi_j(z)\cos z\, d(\cos z)$ is the characteristic of the aperture called by Dubois (4) the "plane angle relation" (FLACHENWINKELVERHALTNIS). As we consider this term too cumbersome, we shall call it the diaphragmatic coefficient.

Assuming that in the first approximation $b(\lambda, z)t_\lambda(z)$ is a linear function of sec z, we shall obtain the following formula for the center of gravity (TN: median value?):

$$\sec \bar{z}_j = \frac{2\int \Phi_j(z)\, d(\cos z)}{\Psi_j} \tag{3}$$

The calculations give for our instrument:

2 mm filter: $\Psi_1 = 0.145$; $\bar{z}_1 = 23.5^\circ$;

8 mm filter: $\Psi_2 = 0.158$; $\bar{z}_2 = 24.0^\circ$.

In the future we shall assume $\bar{z}_1 = \bar{z}_2 = 23.7^\circ$; the remainder of the electromotive forces developed by the thermopile with the lid closed and the instrument open is proportional to the difference of the intensities absorbed by the thermopile in that case and any other.

Considering that the lid is a gray reflector with a coefficient of absorbtion a_o (for carbon black on the shellac varnish, $a_o = 0.96$, according to Albrecht (3), and denoting its temperature by θ we shall have:

$$e_j = p\pi a D S \Psi_j \int [a_o B_o(\lambda) - b(\lambda, \bar{z})] t_{\lambda, j}(\bar{z})\, d\lambda = \frac{1}{m_j} D \int [a_o B_o(\lambda) - b(\lambda, \bar{z})] t_{\lambda, j}(\bar{z})\, d\lambda, \tag{4}$$

where $B(\lambda)$ is the monochromatic intensity of radiation of an absolute black body, and p is the constant quantity for the given thermopile. The quantity

$D\int [a_o b_\theta(\lambda) - b(\lambda, \bar{z})] t_{\lambda j}(\bar{z})\, d\lambda = q_j$, roughly speaking, may be designated as the effective radiation of the lid through the given filter. The quantity $m_j = \frac{1}{p\pi a S \Psi_j}$ is the conversion factor of the instrument, determined at the time of calibration. If m_1 and m_2 are known, then the measurements may be calculated:

$$q_1 = m_1 e_1,\quad q_2 = m_2 e_2 \quad \text{and}\quad q = q_1 - q_2 = D\int [a_o B_o(\lambda) - b(\lambda, \bar{z})] t_\lambda\, d\lambda, \tag{5}$$

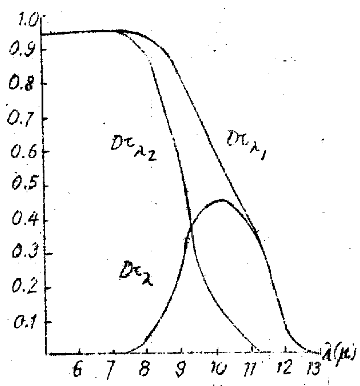
where $t_\lambda = t_{\lambda_1}(\bar{z}) - t_{\lambda_2}(\bar{z})$. The quantity t_λ may be treated as the coefficient of transmission of the imaginary filter equivalent to the pair of filters, so that it would be possible to measure the unknown quantity $q = q_1 - q_2$ with them.

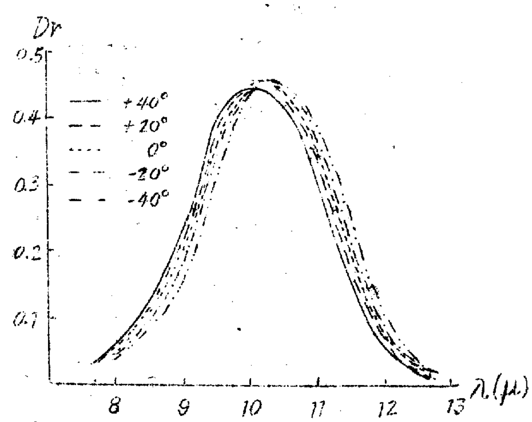
The presence of the two thermopiles allows us to use the instrument as a differential and to obtain q from a single measurement, for which the thermopiles must be adjusted so as to give readings approaching one another. This method, very advantageous for rapid oscillations of radiation in cloudy weather, was not used in the present work, and measurements were always made with either one of the thermopiles.

Knowing q and the temperature of the lid, it is also possible to calculate the intensity of radiation of the atmosphere, the intensity being filtered by such an equivalent filter. Therefore:

$f = D\int b(\lambda,\bar{z})t_\lambda d\lambda = Da_{0}\int B_0(\lambda)t_\lambda d\lambda - q = a_0\beta(0) - q$, where $\beta(\Theta) = D\int B_0(\lambda)t_\lambda d\lambda$. The quantities $t_{\lambda_1}(\bar{z}), t_{\lambda_2}(\bar{z})$ and t_λ were calculated by the coefficients of absorption of fluorite of Reinkober and Kipke (4).

As is shown in figure 3, the equivalent filter will not give the transmission at $\lambda < 7\mu$, where fluorite is almost completely transparent. However, it transmits a considerable quantity of energy the ranges 7 - 8.5μ and 12 - 13μ. In these ranges





water vapor must have a considerable absorption.

Figure 4 shows that the variation of temperature of the filters only causes small deformations of the curve of transparency of the equivalent filter. For working out observations, detailed tables were set up for the functions: $B(t)_{\Theta' = 0°}$ and $\frac{SB(t)}{S\Theta'}$,

where Θ' is the temperature of the filters. It was assumed that D = 0.962. Therefore the absolute value of D does not enter into it, since its deviation from the assumed value is automatically taken into account with the calibrations. The latter were made with the help of a standard radiator for which either melted snow or the bottom of a vessel filmed with warm water was used. The temperature of the water was determined with a thermocouple. If we are to note all quantities relating to calibration and the thermopile, then obviously:

$$m_j = \frac{a_o B_j(0^*) - a^* B_j(t^*)}{e_j^*},$$

where $B_j = D \int B(\lambda) t_{\lambda,j}(\bar{z}) d\lambda$. Therefore it was assumed that for the snow $a^* = 1$ and for the bottom of the vessel $a^* = 0.96$.

The conclusion from formula (4) contains a hidden assumption that opening and shutting the lid of the instrument does not change the temperature of the filters and other parts of the instrument showing an exchange of radiation with the thermopile. However, regardless of the measures taken (regrinding of the filters to fit the sockets of the solid setting and chrome plating the latter), analysis of the readings on the galvanometer with successive opeings and shuttings of the instrument at the time of change convinces us to the contrary. This phenomenon, known in work with radiation-pyrometers (7), is very troublesome for us, since it creates a certain sensitivity on the part of the instrument to radiation of the wave lengths which fluorite does not filter (absorb). If the spectral composition of the radiation were uniform in respect to both measurements and calibration, the phenomenon under consideration would be automatically accounted for.

The above mentioned analysis of successive readings on the galvanometer, which cannot be presented here due to lack of space, convinces us that thanks to the difference in spectral composition it is necessary to introduce a coefficient of reduction equal to 1.02 which may be regarded as a constant, in order to determine the conversion factor directly. Numerous calibrations of the instrument, regardless of the scattering of the various data, show a good correspondence among the averages obtained from snow and from the warm vessel. It is thereby discovered that the instrument has a temperature coefficient of about 0.3 percent degree. Therefore, if $\underline{m}_{1.0}$ and $\underline{m}_{2.0}$ are the constants of the instrument obtained by calibrating and reduced to 0°, then to work out the observations it is necessary to make use of the conversion factors:

$$M_1 = 1.02 m_{1.0}(1 + 0.003\theta) \text{ and } M_2 = 1.02 m_{2.0}(1 + 0.003\theta).$$

Observations were generally made hourly after sunset. A considerable number of readings was taken from each of the filters with the instrument open and shut at intervals of 30 seconds. The opening and shutting of the lid of the instrument was done with the help of a stiff wire rod from a tent in which a vertical galvanometer was placed. All observations, accompanied each time by a determination of the sensitivity of the galvanometer, took from 15 to 20 minutes. In all, 67 good observations were obtained with a cloudless sky. Let us present the extremal values (in units of 10^{-3} cal cm^2 sterradian), together with the corresponding value of the temperature and absolute humidity:

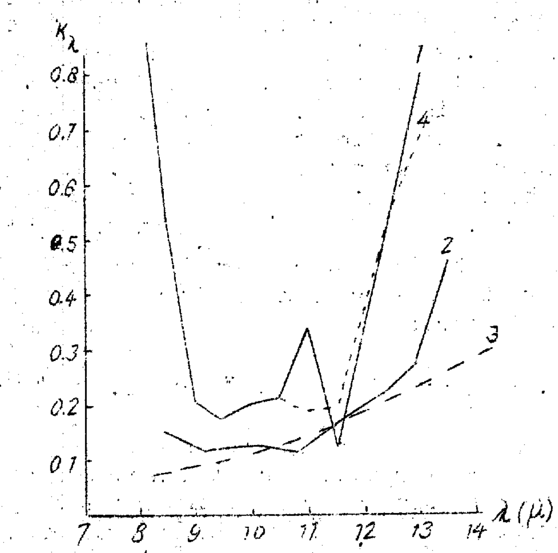
$$f_{max} = 6.5(\tau_0 = 20.5^\circ;\ \varepsilon_0 = 18.7 \text{ millibars}$$

$$f_{min} = 1.0(\tau_0 = 17.3^\circ;\ \varepsilon_0 = 1.2 \text{ millibars}$$

On the analogy of the formulae of Angstrom and Brant, one may grant that in the first approximation $f = B(\tau_0)F(\varepsilon_0)$. If we present the values $\frac{f}{B(\tau_0)}$ for all observations depending upon absolute humidity, a considerable dispersion of various points is evident. However, the dependence of the median values $\frac{f}{B(\tau_0)}$ in each integral of humidity upon ε_0 is characterized by a smooth curve (see figure 6).

In the course of two nights, observations were made every one or two hours. They did not disclose, however, any significant changes in radiation. It is possible that $\frac{f}{B(\tau_0)}$ decreases in the first six hours and increases in the second.

In order to reach some kind of conclusions about the radiation properties of water vapor, experiments were carried out to make the observations correspond with the results of theoretical computations, based on various hypotheses on the magnitude of the mean coefficient of absorption of water vapor in the region of its maximum transparency. The data in figure 5 show how coefficients of absorption of water vapor in this spectral band vary with wave length on the basis of various investigations.



The data of Albrecht (3) are average values. They are obtained by a special reduction of quantities obtained by Hettner under laboratory conditions. These data were frequently used until recently in calculations of the balance of radial energy in the atmosphere. They were presented also in the widely distributed handbook on meteorology by F. Linke (6). On an average they give $K_o = 0.24$ for the spectral band under consideration (curve 1, figure 5).

The data of Adel (curve 2, figure 5) were obtained from changes of solar radiation with a different content of water vapor in the atmosphere, the changes being produced only in regions of continuous absorption, i,e., outside the line. The data of Elsasser (5) (curve 3, figure 5) relate also to continuous absorption. They were obtained by calculation on the assumption that continuous absorption in this part of the spectrum was to be attributed to the edges of rotation bands. (*1) For wave lengths of over 10μ Weber and Randall measured the relative absorptions of various lines, and the method applied completely excluded the background of continuous absorption.

The data they obtained might be roughly considered as average values of relative absorption in various lines, and the averaging was done in a portion of the spectrum, the extent of of which is determined by the practical resolving power of the spectroscope. Therefore, the average relative absorption $\underline{A}$ on any portion of the spectrum $\Delta\lambda$, conditioned by all the lines found in it will be:

$$\bar{A} = \frac{\delta\lambda}{\Delta\lambda} \sum A(\lambda_i),$$

where $\delta\lambda$ is the resolving capacity of the spectroscope in the given section, expressed in units of wave lengths. To this absorption will correspond some average coefficient of absorption:

$$K' = \frac{1}{w} \ln(1-\bar{A}),$$

where $\underline{w}$ is the thickness of the stratum of water vapor (in centimeters of water). Evidently, the full coefficient of absorption will be:

$$K = K_o + K'.$$

(*1) TN: cf Elsasser, Monthly Weather Review, Vol 65, p 324: "Under the conditions realized in the atmosphere, line broadening is almost exclusively due to the perturbation of the radiating molecule by collisions with other gas molecules;" and "The experiments show that with the exception of the gap around 10 μ, the absorption in sufficiently thick layers of water vapor is practically continuous. We may therefore conclude that the main part of the absorption in thick layers takes place in the edges of the lines."

where K_o is the coefficient of continuous absorption.

Using the value $A(\lambda_i)d\lambda$ and w from the work of Weber and Randall, and K_o from the work of Adel, we shall obtain the following mean coefficients of absorption for spectral bands with a range within 0.5μ:

λ...	10.25 ←→ 10.75	10.75 ←→ 11.25	11.25 ←→ 11.75	11.75 ←→ 12.25	12.25 ←→ 12.75	12.75 ←→ 13.25
K'..	0.09	0.06	0.04	0.15	0.33	0.37
K_o..	0.12	0.13	0.16	0.20	0.23	0.31
K...	0.21	0.19	0.20	0.35	0.56	0.68

Thus it would appear that the influence of absorption in the lines cannot be overlooked. The summary coefficients of absorption, as figure 5 (curve 4) shows, correspond fairly well with the data of Albrecht. Nevertheless, the calculations below were made by us for two variants. In the first of these, on the example of Elsasser, we completely disregarded the quantity K', and for continuous absorption we assumed the approximate value $K_o = 0.10$. In the use of the second variant we assumed, with Albrecht $K_o = 0.24$.

Let us consider for a moment the method of calculating the radiation of the atmosphere in relation to the optical properties of our instrument. The radiation of the atmosphere perceptible to our instrument may be considered as the sum of the radiation of water vapor, carbon dioxide, ozone and dust, ie:

$$f = f_{H_2O} + f_{CO_2} + f_{O_3} + f_n .$$

Let $K_\lambda(h)$ be the coefficient of absorption of the matter under consideration, and $p_k(h)$ the density of this matter. Then, each of the three components of radiation can be calculated by the formula:

$$f_k = D\int_{\lambda=0}^{\infty}\int_{\xi_k=0}^{\chi_k} B_t(\lambda)t(\lambda)e^{-\xi}\,d\xi_k\,d\lambda = \int_0^{\chi_k}\sum\left(D\int_{\lambda_i}^{\lambda_j} B_t(\lambda)t(\lambda)d\lambda\right)e^{-\bar{\xi}}d\bar{\xi}_k, \qquad (7)$$

where $\xi = \sec\bar{z}\int^{h} K_k(h)p_k(h)dh$; $\chi_k = \sec\bar{z}\int_0^{\infty} k_h(h)p_k(h)dh$;

$\xi = \sum\xi_k$; and $\bar{\xi}_k$ and $\bar{\xi}$ are the mean values of ξ_k and ξ –for spectral band $\lambda_i - \lambda_j$. If the individual components $(f_{H_2O}, f_{CO_2}, f_{O_3})$ are calculated, the formula undergoes considerable reformation and simplification. We, however, can not stop here for these reformations due to lack of space.

We shall limit ourselves to only the following observations. In calculating f_{H_2O}, it was assumed with Schmidt, that the coefficient of absorption of water vapor is proportional to the square root of the pressure and, conversely, proportional to the fourth root of the absolute temperature. Absorption in the weak band 9.1 - 10.9μ of carbon dioxide was calculated by use of a theory of absorption in spectral bands which Schnaidt had worked out. Therefore, in agreement with the theory and in distinction from the case of water vapor, it was considered that the coefficient of absorption of carbon dioxide is proportional to the pressure, and conversely, proportional to the square root of the absolute temperature. Data on the absorption of ozone in the infra-red band 9.1 - 10.0μ are unknown to us. Therefore, on the basis of figure 7 in the book of Getts *(TN: probably Goetz), Atmospheric Ozone, we assumed purely tentatively that the range of the deep portion of this band equals 0.6μ and has an absolute transparency (for the whole mass of atmospheric ozone.)

The content and optical properties of dust are so indefinite that we prefer to refrain even from a rough calculation, and so determine the radiation for the corresponding component from comparison of calculated and observed quantities. In table 2 are shown $f_{H_2O} + f_{CO_2} + f_{O_3}$ and their sum and relations

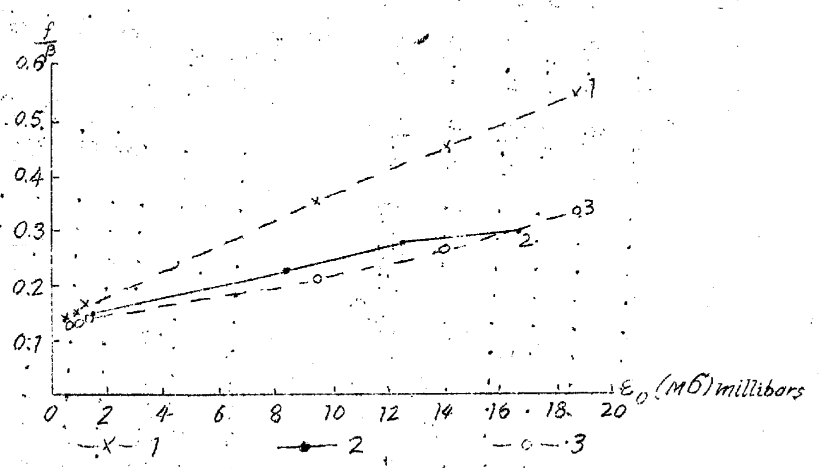
$\frac{f_{H_2O} + f_{CO_2} + f_{O_3}}{B(t_o)}$, calculated on the following assumptions on the stratification and content of the atmosphere:

1. Distribution of water vapor follows the known formula of KHERGEZEL *(TN: probably Hergesell)

2. The gradient of the temperature in the lower levels of the atmosphere for a distance of one kilometer equals 0, but above, as far as the stratosphere, it equals 0.6°C per 100 meters.

3. The content of O_3 at different heights has the values assumed by Vassy.

The calculations are carried out for two values of the temperature and three values of relative humidity at the surface of the earth; and for both assumed values, indicated above, of the coefficient of absorption of water vapor.



$(f_{H_2O} + f_{CO_2} + f_{O_3})^{\beta}$ is determined by calculation; in 1 - k_o = 0.10; in 2 - k_o = 0.24; and in 3 - $\frac{f}{\beta}$ is determined by observation.

Table 2

	$t_o = -20^o$						$t_o = +20^o$					
u_o(%)	40		60		80		40		60		80	
K_o	0.10	0.24	0.10	0.24	0.10	0.24	0.10	0.24	0.10	0.24	0.10	0.24
f_{H_2O}	0.06	0.15	0.08	0.22	0.12	0.30	1.98	4.44	2.90	6.10	4.05	7.51
f_{CO_2}	0.06	0.06	0.06	0.06	0.06	0.06	0.13	0.13	0.13	0.13	0.13	0.13
f_{O_3}	0.78	0.78	0.78	0.76	0.78	0.75	1.06	0.87	0.97	0.74	0.93	0.64
$f_{H_2O} + f_{CO_2} + f_{O_3}$	0.90	0.99	0.92	1.04	0.96	1.11	3.17	5.44	4.00	6.97	5.11	8.28
$(f_{H_2O} + f_{CO_2} + f_{O_3}):B(t_o)$	0.13	0.14	0.13	0.15	0.14	0.16	0.21	0.35	0.26	0.45	0.33	0.54

Table 3

Air Masses	13 Aug кПВ		26 Aug трАВ		9 Sep кТВ	
t_o(°C)	14.4		7.9		23.5	
ε_o(millibars)	12.8		7.6		15.7	
K_o	0.10	0.24	0.10	0.24	0.10	0.24
f_{H_2O}	2.90	5.90	1.53	3.33	3.68	7.40
f_{CO_2}	0.12	0.12	0.10	0.10	0.14	0.14
f_{O_3}	0.89	0.66	0.90	0.75	0.84	0.63
f_π	0.42	0.42	0.37	0.37	0.47	0.47
f calculated	4.33	7.10	2.90	4.55	5.13	8.64
f observed	4.8		3.3		4.6	
f calculated-f observed	-0.5	+2.3	-0.4	+1.3	+0.5	+4.0

As can be seen graphically from figure 6, the second variant ($K_0 = 0.24$) will give greatly increased values of atmospheric radiation, as well as an increased dependence of this quantity upon absolute humidity. Conversely, the first variant ($K_0 = 0.10$) will give a dependence, agreeing with the observations, upon the absolute humidity and in general will give only slightly lowered absolute values. The latter circumstance naturally includes the influence of dust. In this case it is necessary to assume $\frac{f'_\pi}{B(t_0)} = 0.03$.

Three times (13 and 26 Aug and 9 Sep 1944) at the time of measurement of radiation, special ascents for radiosonde were made. Therefore, in given cases, the radiation of the atmosphere may be calculated more accurately by using the observed values of the temperature and humidity at various heights. Comparison of observed and calculated values of radiation is made in table 3. Therefore, in agreement with the above, it is assumed $f_\pi = 0.03$ $B(t_0)$, and the radiations of other components of the atmosphere are increased to 0.97 (proposed coefficient of filtration of dust.)

The data presented here completely bear out the conclusions reached. Small discrepancies between the calculations on the first variant and the observed quantities can be entirely explained by incalculable variations in the content of the dust, ozone and carbon dioxide.

Thus, we may formulate the following 2 conclusions:

1. The mean coefficient of absorption of water vapor in the region of its maximum transparency is about 0.10 (cm $H_2O)^{-1}$.
2. The intensity of radiation of dust under the conditions of the Kuybyshev Observatory consists of about 37 percent of the radiation of a black body at earth temperature.

The uselessness of the data of Albrecht for coefficients of absorption of water vapor in the investigated region is not in itself strange, since they were obtained by a formal mathematical method. However, the value $K_0 = 0.10$, corresponding to only one continuous absorption, seems too small. The cause of this discrepancy remains obscure.

The observations, forming the basis of this work, were completed in their main form by the technicians of the actinometry group of the Kuybyshev geophysical observatory, and the radiosonde by A. S. Steparoviy, specialist of the aerology group. The director of the Observatory, T. G. Moskvin, was very helpful at all stages of the work.

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II. ON THE DISTRIBUTION OF ATMOSPHERIC PRESSURE OVER THE CENTRAL ARCTIC

by

B. L. Dzerdzeyevskiy

In number 12 of the jounal Arctic Problems, 1940, appeared some monthly charts of isobars in the Arctic Circle, drawn up by D. B. Karelin (4). The work was based on data from 75 stations, 49 of which were located north of 60 degrees north latitude and 33 within the Arctic Circle. Only those stations were used which have a full 8-year series of observations. This principle, however, was not rigidly applied, since it was possible to fill various gaps by interpreting available evidence, and data of the stations North Pole and Sedova are used. Their observations, it seems, were also cited for the 8-year period.

There is a great practical possibility of evolving more accurate systems of diagraming the distribution of pressure in high latitudes; and the theoretical significance of such systems, as well as the progress of Soviet work in the arctic, the organization of the vast network of the polar stations, the many expeditions, and finally, the observations in the very center of the arctic, all indicate a possibility in this direction.

All these materials were completely exploited by myself in working out the observations of the station North Pole (2,3).

In comparing my maps with those of D. B. Karelin, I was obliged to note a marked discrepancy between them. The great importance of the problem suggests analysis of the causes of such a discrepancy, which I shall try to make. Since both D. B. Karelin and I compared our maps with the analogous maps of F. Baur (1), I shall refer to them here.

On the maps of Baur, Sverdrup (6), and myself, isobars are drawn for every 5 millibars; and on D. B. Karelin's, every 1-2 millibars. This creates an impression of great accuracy of detail in reflecting the distribution of pressure. But this accuracy is only apparent. Diminishing the integrals between the isobars can show new details only if there is an extraordinarily close network of stations to verify and prove such accuracy. In the absence of such detailed data to start with, the "detailing" is reduced at once to drawing a large number of meaningless parallel isolines. The maps of D. B. Karelin are a prime example of this (figure 1).

He devotes sufficient attention to the concurrence of the general configuration of lines at the peripheral parts of all three series of maps (especially in the winter half of the year) and their divergence in the central parts of the region (especially in the summer half of the year). This very area and season are most interesting.

Baur's maps were based on data gathered from 1850 to 1927 in the high latitudes. The duration of observations at each point was different. More important, the series of simultaneous observations was very small. Consequently, Baur's application of such variegated data to one period is rather questionable and is confirmed, up to a certain point, only by the absence of other possibilities. For the central arctic, Baur had at his disposal only the observations of the "Fram," and the distribution of pressure in this most interesting area remained the least detailed and verified.

The data of recent years may be used in two ways:

1. Adding them to the already existing series of data and constructing maps based on the series thus amplified.
2. Proceeding only on the basis of the uniform data of recent years.

The short duration of the period, it is true, makes the climatological authenticity of the conclusions questionable; but it is quite evident that even the first variant under these conditions will not give this assurance.

In fact, what degree of reality will the short-time observations of the station North Pole and Sedova have after they are applied to the long-term period, obtained in turn by collation of short and nonsuccessive periods from different stations? Undoubtedly, such an operation vitiates the value of new data.

Accordingly, in my work I made no collations with the long-term period, using only simultaneous observations. In conformity with the general problem, the years with observations in the central arctic were taken---1937-1939. D. B. Karelin shares this point of view, but he does not achieve it practically; rejecting the application of recent data to the long-term period, he nevertheless includes data of the station North Pole and Sedova in the 8-year series, evidently applying them to this series.

I consider that the possibilities permitted by new observations are to a great extent vitiated in this way.

In any case, let us determine the different duration of the periods taken by D. B. Karelin and myself. Could it be that the difference between them noted above can be explained just by this? Such an explanation is unsatisfactory. A little later, I shall show why. For the present, I am demonstrating the fact that, in any areas, and particularly in the central arctic, one cannot draw isobars, so to speak, mechanically, without a definite, logical, guiding concept and without means of checking the resulting conclusions with other material. In this case, the system of the general circulation of the atmosphere must be the leading concept, as well as the dynamic connections between its basic members, established and checked in the light of good observations (including aerological) from other places on the globe and by the methods of checking called synoptic analysis. Without this system of diagraming, the isobars will be indecipherable and inconsistent. Strictly speaking, it will not be possible, as a rule, to formulate certain noteworthy systems.

I shall use as an illustration for the above a series of D. B. Karelin's maps for various months of the summer half of the year.

How, for example, can the July map (figure 2), with three adjacent but isolated anticyclones located on different stratal surfaces, be dynamically explained? Or how can the huge, many-centered Asiatic depression outside have equal status with them (Ed N: apparently refers to "anticyclones" above)? Having adopted such a system, already apart from any other, we shall have to recognize the existance, during this month, of a steady, southerly flow across Alaska and Chukotka, as steady as the northerly and northeasterly flow to the Taymyr Peninsula. The observations certainly do not support this.

In August (figure 3), the center of the polar anticyclone (p>1,012 millibars) is located between Wrangel Island and the coast of Chukotka, and isobar 1,011 is formed without reaching 120 degrees west longitude. The second anticyclone occupies all Alaska and Canada, and its center is located in subtropical latitudes; but both these anticyclones are denoted by one isobar, forming an inordinately extended crest of raised pressure

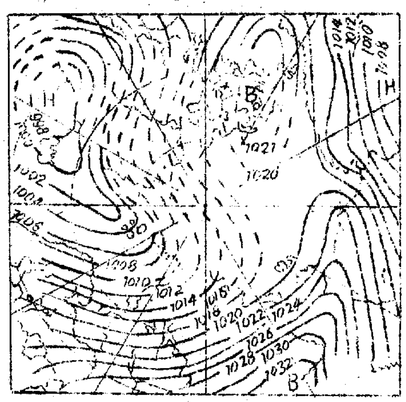

Figure 1

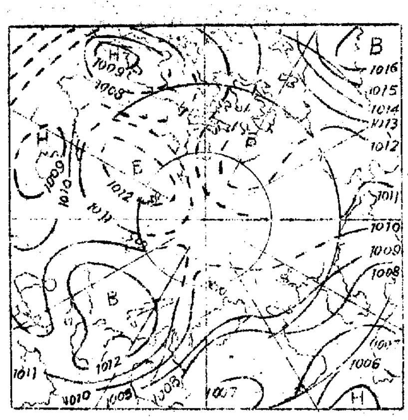

Figure 2

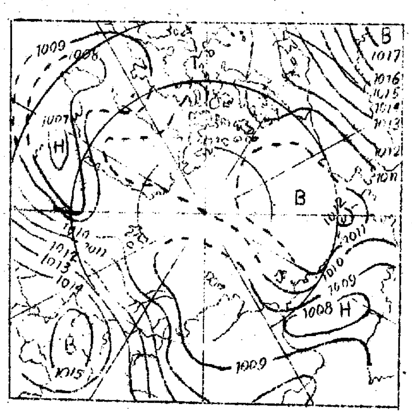

Figure 3

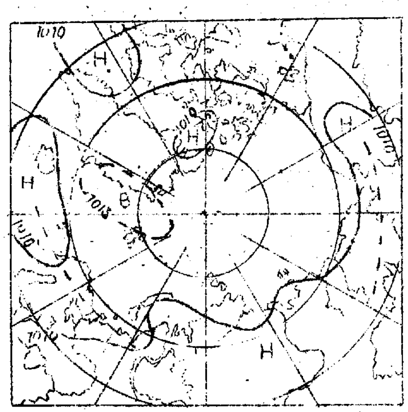

Figure 4

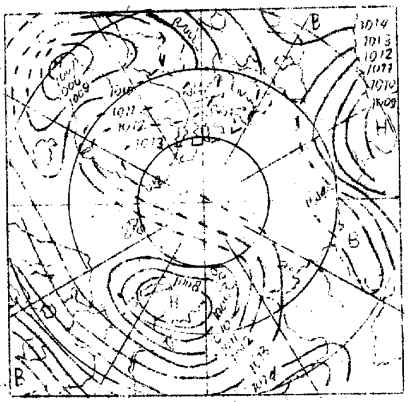

Figure 5

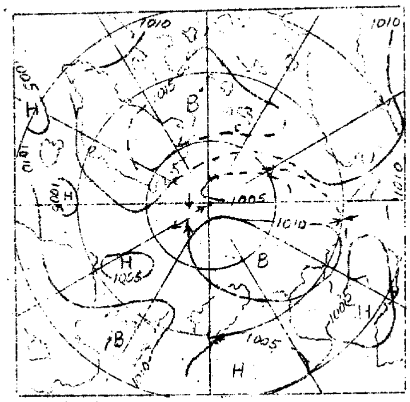

Figure 6

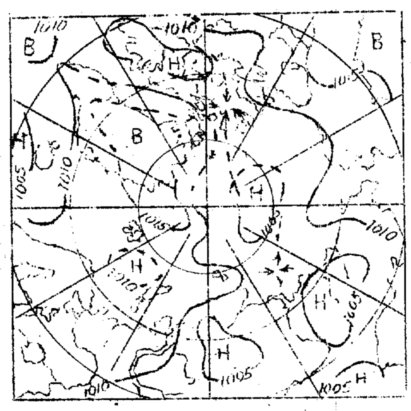

Figure 7

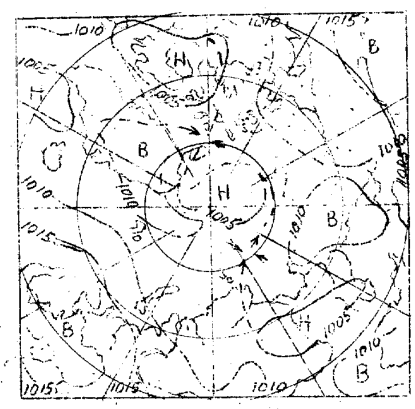

Figure 8

to Greenland. Everywhere in the central arctic there will remain a field with no gradient actually permitting one, to set up any process or system here. On Baur's map for this month (figure 4), the field is also without a gradient; but certain isolines have not been drawn on it, even with dots, and it appears to emphasize the desire to not enter into the field of unsupported suppositions.

In September (figure 5), the system of the preceding month is repeated in still more pronounced form by D. B. Karelin. Thereby, an extremely indeterminate map is created, unable to bear the test of its dynamic basis.

Thus, by drawing isobars mechanically, refering only to the value of the pressure at various points, it is possible to obtain ten more baric systems, not completely similar. And no one of them will have preference over the others.

However, the guiding concept in D. B. Karelin's maps evidently was the necessary existence of a polar anticyclone during the whole year. This also led to unjustified systems, but with, at least, a similarity to the region of raised pressure. If such a concept could have been justified in Baur's case (although he used it very carefully), it is more difficult to agree with now.

As contemporary materials have shown, the number of days in the summer months with cyclonic activity in the central arctic is equal to, or more than, the number of days with anticyclones. Synoptic analysis of cyclones has shown that their structure and phases of development correspond to our ideas of the development of analogous processes in subpolar and temperate latitudes. Therefore, it was possible to establish the predominance in summer of young wave cyclones in the central arctic, the existence of steady forming fronts, and consequently, the existence of circulation systems, guaranteeing the development of the inevitable fields of deformation.

Not stopping here with the series of extra considerations and data used in the construction of the maps, I shall refer to a few of them (figures 6-8). The characteristic peculiarity of the three summer months (June, July and August) is the prevailing depression in the polar regions. Returning to the maps of D. B. Karelin and especially those of Baur, it is easy to determine that the place for this depression is located on the maps of both, and it makes the system on the whole map more logical and authentic.

There is indicated everywhere on my maps, in the months of the summer half of the year (May-October), the position of the frontal zones and the hyperbolic points generating their fields of deformation. The seasonal displacement of the arctic front has been determined---from the usual position for winter (November-April) and neighboring months (May and October) near the latitude position to the summer one near the meridional position-from the continents to the polar depression.

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The total lack of aerological observations in the central arctic prevents giving exhaustive explanations for the determined facts. Now we must confine ourselves to the hypotheses alone. Remembering that the amount of radiation in the arctic in summer is larger than anywhere on earth, the formation of the polar depression may be connected with it. This is confirmed by the slow degeneration of the warm air masses here. Depressions of thermal character do not usually connect with fronts. In the present case, we evidently have a combination of thermal and dynamic factors, reinforcing each other. The process is begun chiefly by the arrival in the central arctic of a deep occludent cyclone. The deepening here, under the influence of the peculiar conditions, makes the depressions strengthen the flow and regeneration of fronts of occlusions. In this way, the dynamic and thermal conditions, reinforcing each other, bring about a well developed process.

In order to be convinced of the authenticity of the new systems, some checks were devised. In particular, on the synoptic maps of the northern hemisphere, a number of days were noted in which each of the stations was in the cyclonic circulation. Averages of each month for three summers were put on the map, and isolines were drawn in accordance with them. These maps well supported the intensive occurence of cyclones in the polar areas, as shown in the distribution of average pressure, the correspondence of frontal zones with the axes of the region of maximum cyclonic occurrence agreeing with it. (TN: ie, the pressure.) The seasonal symmetry of the trajectory of the hyperbolic point of the field of deformation forming the arctic front, its comparison with the analogous system of H. Sverdrup (6) for trajectories of instrusions into the arctic atmosphere in various seasons, and the course of the quantity of precipitations, all supported the conclusions harmoniously.

It remained to solve the problem of what stability the obtained systems have, and whether they were merely conditioned by the peculiarities of the 3-year period studied. Can it be that the discrepancies between them and the maps of D. B. Karelin are due to the fact that he took a longer period?

For checking the conclusions from this point of view, aside from the comparison with the data of H. Sverdrup, the ample and comprehensive work of E. S. Lir (5) was used. Using data covering 35 years, it established weather types determined by the dispostion of flow, opposite in their quantities and direction. Two such streams bring about the formation of a front between them.

Mapping the position of the rectilinear flow of E. S. Lir and the position of my own frontal zones, I obtained a satisfactory correspondence for all months. In this way, the comparison of results based on material of 3-and 35-year periods bears out the great stability of the position of the fronts, and consequently, of the circulation systems which condition them. Therefore, the application of actually homogeneous material alone is justified, even if over a short period of time, since it is a period, with a basic advantage, observations in the central arctic, along with the extended network of polar stations permitting one to process the material with synoptic analyses.

Consequently, the discrepancies between my maps and those of D. B. Karelin are not due to the difference in the length of the periods. I have tried to demonstrate the fundamental nature of these descrepancies.

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(TN: Asterisks indicate a Russian original.)

III. A STORM OCCURRENCE OVER THE CENTRAL REGIONS OF THE EUROPEAN TERRITORY OF THE USSR

by

I. G. Pchelko

On the night of 23 Sep 1943 and during the day of the 24th, stormy weather with an exceptionally strong wind was observed over the central regions of the European Territory of the USSR. The oblasts of Smolensk, Kalininsk, Moscow, Tul'sk, Ryazansk, Ivanovsk and Gor'kovsk, chiefly, were buffetted by strong winds. In these provinces the wind reached hurricane proportions and caused great damage to the national economy.

Unfortunately, adequate investigation of the phenomenon cannot be carried out because the synoptic processes giving rise to the strong winds developed in great part over territory that was not observed at the time, and consequently, we do not have at our disposal the meteorological and aerological data necessary for investigation. The whole analysis of the general synoptic situation, with the advantage of even very limited aerological data, is highly interesting, and the results of this analysis will be useful not only for working with an abridged map, but also with a fairly complete map and for any other region.

After a prolonged period of anticyclonic weather which enveloped a considerable part of the European territory of the USSR from 17-18 Sep 1943, a period of cyclonic activity set in, developing in a direction from Iceland toward the Barents Sea. On 20 Sep the intrusion of a cold, maritime, arctic air mass was noted in the rear of the cyclonic region, occupying the area of the Barents and Norwegian Seas. At 0700, 21 Sep, in view of the strong northern weather, the arctic intrusion, enveloped the whole territory of the Norwegian Sea, the Scandinavian peninsula, and the northern areas of England. The speed of this flow at a height corresponding to 700 millibars was 50-60 kilometers per hour, and at a height corresponding to 300 millibars it was 70-80 kilometers per hour. The position of the arctic front on the morning of 21 September is indicated on figure 1 by two thick, stippled lines. At 1900, 22 September (figure 1) the arctic front advanced from the cyclone over the Barents Sea across Lake Onezhskoye, Polotsk, and far to the southwest, presumably across the southern districts of Germany. To the east of the principle cold front, at a distance of approximately 200 kilometers, a bent branch of polar front occlusion was noted.

The synoptic map for 1900, 22 September (figure 1) should be of the greatest interest to us since strong winds over the central regions of the European territory of the USSR began just 24 hours after the time of this map, and therefore with a good prediction a warning of the danger could be given on the basis of this map 16 to 18 hours before its onset. What are the peculiarities of the synoptic processes depicted by this map, and

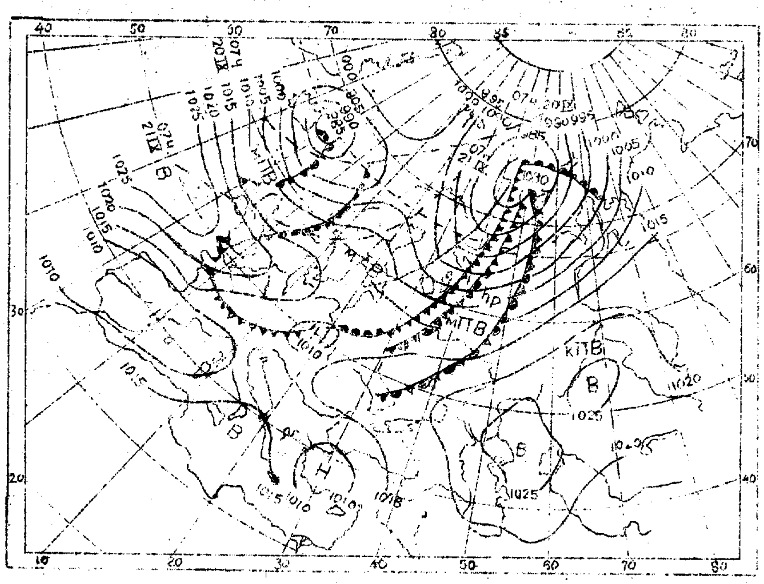

Figure 1

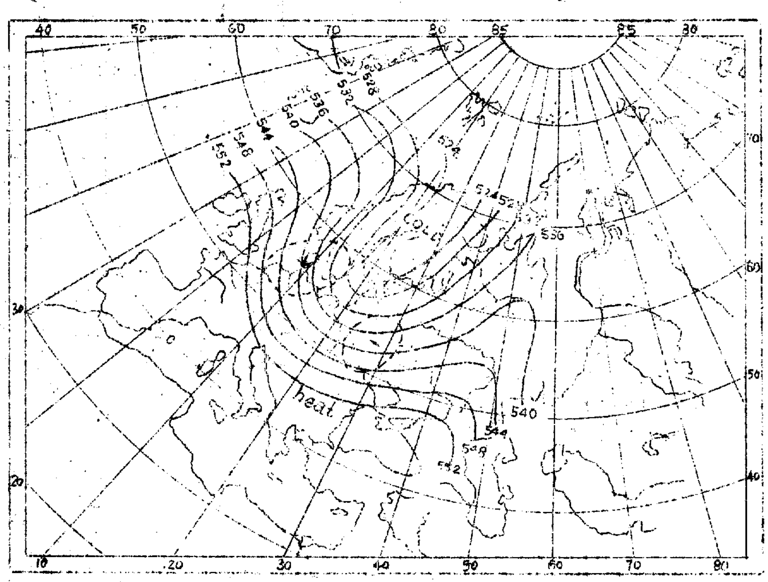

Figure 2

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what details are worthy of serious attention? It is evident that an ominous development was taking place in the atmosphere at this time, and it should certainly have been announced.

The arctic front over reported territory was very weakly marked on the surface of the earth; it was discovered by only small breaks in the field of baric tendencies, a slight rain was recorded in places in the frontal zone, winds on this and the other side of the front were weak and almost in one and the same direction, and the temperatures of the air masses along the front differed in all only by 2-3 degrees. All this indicated synoptically only that the frontal zone in question is hardly the main front, and that this front evidently dissipates and therefore lacks an adequate supply of energy. It was assumed that this front, over unobserved territory, was located in a quasi-stationary position, in a gradientless baric field.

It is evident that wave formation was already taking place on this part of the front. Such a disturbance could have been formed by a rising, normal component of atmospheric currents at the main front, which, in their turn, were a consequence of the presence of a secondary cold front and a trough connected with it, located somewhat north of the main front (figure 1). The wave disturbance, denoted by a closed isobar of 1010 millibars, was formed over the territory of Poland.

Having set up a synoptic background, the question arises: if there is a wave disturbance, will it become developed and how intensively? Analysis of terrestrial data on the reported territories, as was stated above, leads to negative conclusions. Aerological data, unfortunately very limited, should be more significant, if these data are timely and correctly analyzed.

On a map of relative topography a frontal zone of 500/1000 millibars (figure 2) altitude between the cold area over northern Europe and the warm area over southern Europe corresponds to the position of the arctic front. To judge by the thickness of the isohyp, the temperature contrast in this zone averages 24 dekabars per 1,000 kilometers; that is, very close to a critical value, determined for active frontal zones by the investigations of Pogosyan and Taborovskiy. The difference in equivalent potential temperatures between the former maritime polar atmosphere and the maritime arctic atmosphere was about 8 degrees which also, according to Bergeron, is rather close to a critical value for the main frontal zone. But above all it is necessary to consider the change in the speed of the wind in the given frontal zone as characteristic. The speed of the wind on the ground along the frontal zone on the morning of 22 September did not exceed 15-20 kilometers per hour, and at a height corresponding to 700 millibars, by the observations in Leningrad and the Smolensk-Velikiye Luki area, it reached 60 kilometers per hour. From the morning to the evening of 22 September, the wind at the earth's surface remained weak as before, while it strengthened to 80-90 kilometers per hour during this period at a height corresponding to 700 millibars. Such a sharp increase in wind velocity at the time and at that altitude was clearly connected with the increase of the temperature contrast at the frontal zone, and consequently it indicated an accentuation of the zone and a possibility of an active cyclogenesis on the arctic front. According to Troytskiy (3), as is known, this is an indication of a cyclone issuing from the southwest. And finally, the supposed wave disturbance over the territory of Poland was located near the delta of the frontal zone, which, in accordance with the investigations of Pogosyan and Taborovskiy, points to the possibility of an intensive development of this disturbance.

At 0700, 23 September, for a period of 12 hours, the position of the newly formed cyclonic disturbance was more definite, although a large part of it was still over unrecorded territory. The center of the disturbance was assumed to be in the region of Minsk with a pressure somewhat below 1,000 millibars. The zone of precipitations, connected with the warm branch of the arctic front, enveloped all of Kalininsk and parts of Smolensk and Moscow Oblasts. The negative baric tendencies reached 3-4 millibars, and a sharp decrease of pressure and a second area of heavy precipitations was noted not only against the warm front, but also south of it in the warm sector. The last circumstance obliged one to assume the existence in the warm sector of some other frontal division. It is difficult to say what the nature of such a front was. Possibly, this was the residue of the old depression over western Europe and the disturbance we are considering is the result of the regeneration of such a depression. But such a supposition is difficult to reconcile with the history of the development of the processes in the west on the preceding day. One might also suppose that this front is the end of the bent part of an occlusion of the cold front type, the part connected with the cyclone over the Barents Sea. In any event, the existence of a front ahead of the main cold front was corroborated when the process shifted to recorded territory. The winds on the earth's surface in the area of the developed cyclone were still very weak and did not exceed 2-3 balls (TN "balls" are numbers in a scale such as the Beaufort wind scale). Probably, the application of Gilbert's well-known rule would be very pertinent here: "If a drop in pressure in a cyclonic area is very marked but the winds are weak, the cyclone may turn into a very strong rotor."

At 1900, 23 September, the center of the cyclone was already in the Mozhaysk area with a pressure of about 986 millibars (figure 3). Thus, the disturbance was deepened for 24 hours to approximately 22 millibars and was displaced during this period for approximately 1500 kilometers, advancing at an average of about 60 kilometers per hour. The comparison of the average speed of the cyclone at the earth's surface with the average speeds of the main current at the higher levels indicates that in relation to the speed at a height corresponding to 700 millibars, the speed of the cyclone at the earth's surface was 0.75, but in relation to the speed at a height corresponding to 500 millibars, about 0.60. The cyclone became a strong whirlwind with very large horizontal baric gradients. The latter reach 5 millibars per degree of the meridian in the warm sector of the cyclone and give rise here to southern and southwest winds up to 7-8 balls. But the largest gradients were created in the rear of the cyclone, where they reached ten millibars per degree of the meridian and

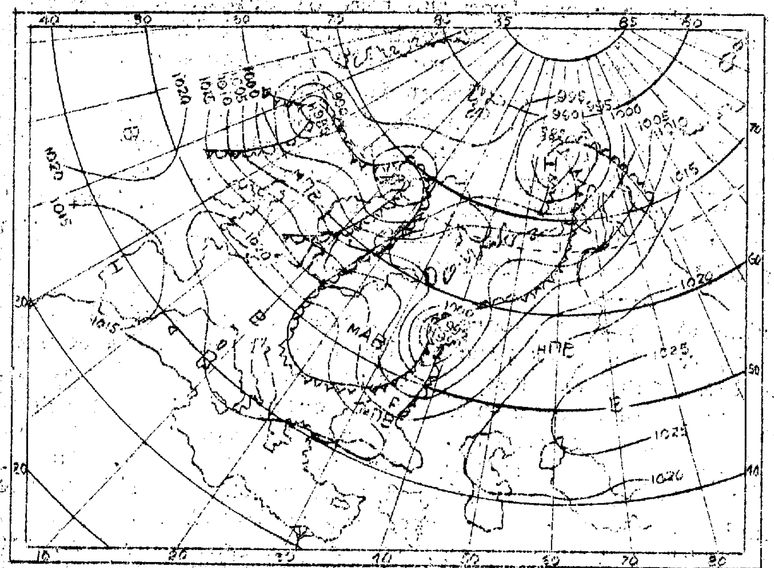

Figure 3

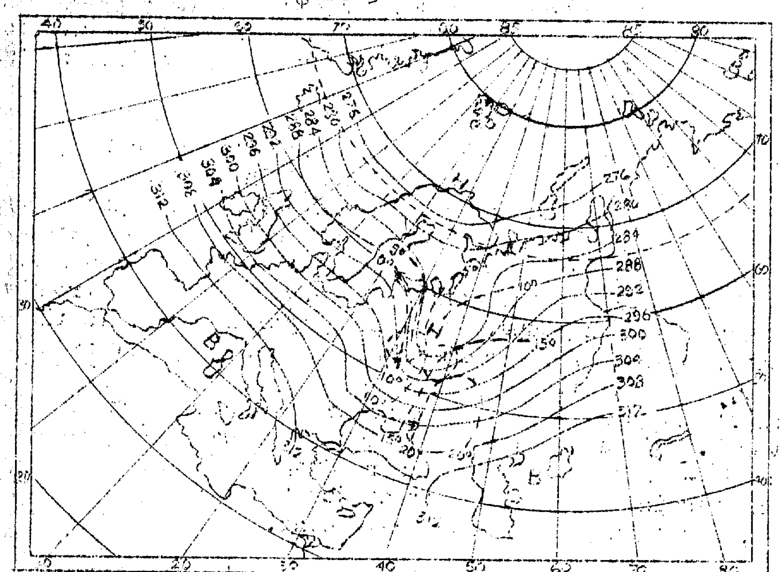

Figure 4

conditioned the winds of the north quarter to 10-11 balls (over Kalininsk and Smolensk Oblasts). The disturbance at this time reached the stage of a young and ideal cyclone. Its deepening continued until midnight on 24 September, when the pressure in the center went down to 983 millibars. Thus, the general deepening of the cyclone from the time of origin to the stage of highest development was approximately 25 millibars. The baric tendencies both in the area of decreasing and in the area of growth had at this time almost identical absolute values; that is, about ± 5.0 millibars. In the following hours the growth began to exceed the decrease and consequently the cyclone was already beginning to form. The force of the wind reached its highest value about this time, and the northwest wind was at its strongest at the rear of the cyclone, where it coincided with the area of the greatest increase in pressure.

During the next 30 hours the occluding and gradual formation of the cyclone took place. The wind gradually weakened at the same time. During these 30 hours the cyclone shifted from a fairly large northern component in accordance with the changed direction of the main current. The average speed of the cyclone decreased to 40 kilometers per hour over this period. The speed of the main current in the upper levels of the atmosphere also diminished. At a height corresponding to 700 millibars it averaged 60 kilometers an hour, and at a height of 500 millibars--about 70 kilometers an hour. Thus, in this period of the cyclone's development, the ratio between the speed of the cyclone on the ground and the speed of the wind in the upper levels was about 0.7 relative to the speed at a height corresponding to 700 millibars and 0.6 relative to the speed at a height corresponding to 500 millibars.

Let us stop to consider some very important peculiarities of the cyclonic disturbance in question. The first of these relates to the distribution of baric tendencies. As indicated above, with the approach of the cyclone to the central areas of the European territory of the USSR, negative baric tendencies occurred both ahead of the warm front and in the warm front, as usually happens with young, developing cyclones, but in the warm sector, ahead of the cold front. On the map for 1900, 23 September, when the cyclone had almost reached its highest development the negative baric tendencies in the warm sector were over 7.0 millibars, while ahead of the warm front they did not exceed 3-4 millibars. The sharp drop in pressure changed after the passing of the cold front due to a strong increase in pressure, accompanied by a hurricane wind.

A drop in pressure in the warm sector is always observed in sharply deepening cyclones. This phenomenon is generally connected with the upward displacement of the atmosphere of the warm sector and with its decreasing outflow from the area of the cyclone; that is, with divergence of the upper currents in the given region. In our case this dynamic effect is plainly very essential. On the map with topography corresponding to 700 milibars (figure 4) it can be determined that the divergence of the upper currents begins in the warm sector ahead of the cold front with very high speeds (80-90 kilometers per hour). This divergence can be judged more clearly by the position of the isotherms at the surface of the earth. The latter are drawn in figure 4 with broken lines, indicating the striking convergence of the isotherms with the cold atmosphere beyond the cold front and their still more striking divergence in front of the cold front in the warm sector. Thus, following the still valid rules of Sherkhag that it is possible up to a certain point to judge the convergence and divergence of upper currents by the earth's surface, we can explain the sharp drop in pressure which takes place in the warm sector ahead of the cold front.

The warm front, in contrast with the cold front, was located in a region of decidedly weaker winds; this was also pretty noticeable for the same front in the lower layers, and evidently, the convergence of friction in the lower layers in great part compensated for the divergence above. Consequently, the effect of the divergence of the upper currents was significantly weak here, and therefore, there was no such extreme drop in pressure ahead of the warm front which is usually noted with young and developing cyclones.

The second peculiarity of the cyclone appeared in the following: over the whole period of development and activity of the cyclone from the southwest to the northeast, its warm front did not show any marked progressive displacement. In other words, the amplitude of the wave was not increased, and consequently, the warm atmosphere did not advance northward into the area occupied by the cold arctic atmosphere. As a result the path of the center of the cyclone did not deviate to the left by 10-30 degrees from the general direction of the frontal zone but coincided almost exactly with the position of the arctic front ahead of the center.

The position of the arctic front for the period 22, 23 and 24 September is shown in figure 5, as well as the path of the cyclone from 22 to 25 September. It is evident from this diagram that the movement of the cyclone in the course of the 24-hour period nearest the selected initial moment coincided with the position of the warm front at the given moment, and the path of the cyclone over the 24-hour period was roughly equivalent to the length of the whole area of the warm front. It can be said that the direction of the advance of the cyclone coincided with a straight line joining the center of the area of increased pressure behind the cold front with the center of the area of negative tendencies in front of it.

The absence of any marked progressive displacement of the warm front of the cyclone may be explained by the absence of normal components of motion in the cold atmosphere ahead of the front, both in the upper and lower strata. Thus the warm branch of the arctic front remained in a quasi-stationary position, and consequently, the warm atmosphere did not rise in the area occupied by the cold arctic air. As far as the path of the cyclone is concerned, it was already established by V. A. Bugayev (1) in 1936 that cyclones regenerating on arctic fronts (for cases in

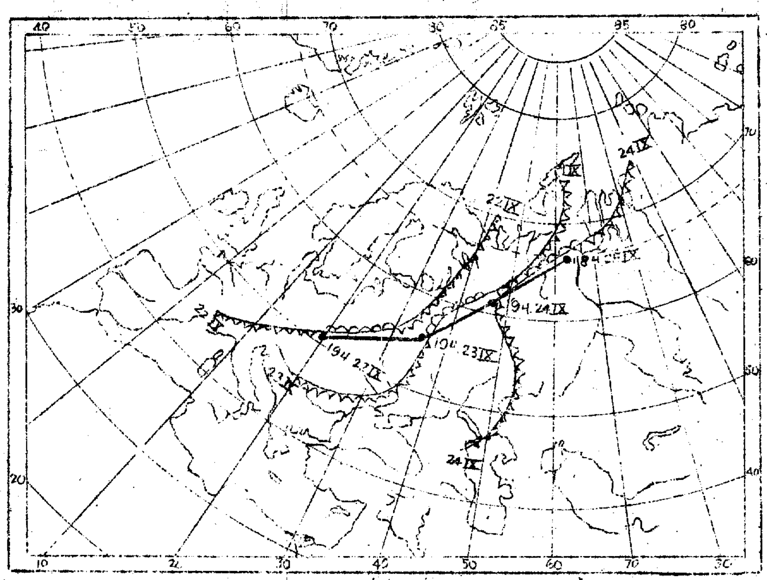

Figure 5

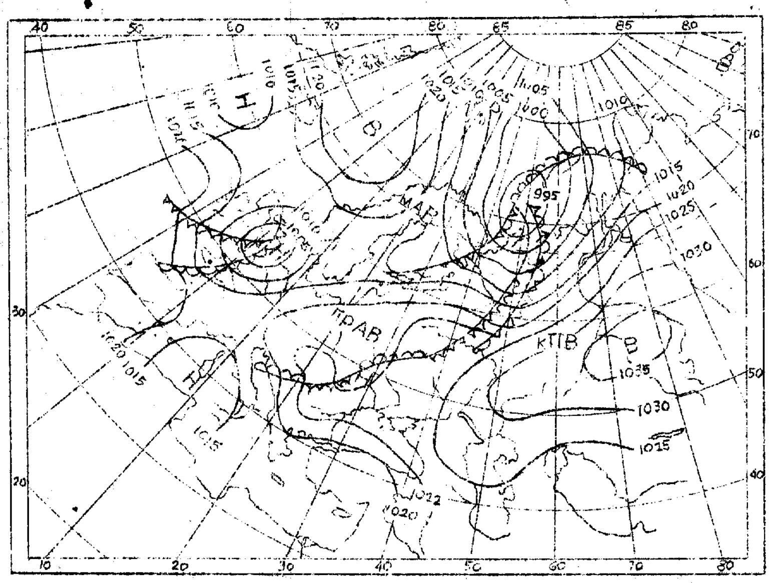

Figure 6

which the arctic atmosphere did not extend so far to the south) advance on a path almost coinciding with the position of the arctic front ahead, tending to bend from the south around the mass of arctic air. It is accordingly referred to the statement of Shinze: "In the cold time of the year, stationary and even vertically undeveloped cold masses are a great obstacle to the penetration of warm masses, since the expenditure of a particularly large amount of energy is always required to displace the stationary cold masses." Such an explanation can hardly be considered convincing in our case, if only because we have to do with a process of the transitional or warm time of the year, and the cold air in the forward part of the disturbance did not constitute such a mass as to present a great obstacle to the penetration of the warm air. The cause of the said shift of the cyclone and its warm front is evidently to be found in the structure of the upper currents, and in particular, in the location of the field of convergence and divergence of the current.

In conclusion, let us make on comparison. A few days after the case considered here, very similar synoptic processes were repeated. Over the European territory occurred a repetition of the cold, maritime, arctic air mass from the same region - that is, from the direction of Spitzbergen and Greenland. On 25 September the arctic frontadvanced over the northern regions of England, and at 0700, 28 September it occupied the position Kanin Cape-Moscow-Kiev and beyond, almost to the northern shore of the Adriatic Sea. And as in the first case, this front was so weakly manifested over our territory that the synoptic operator on duty generally doubted its existence and marked the frontal zone only with a dotted line. It was supposed that over the southwestern areas of the European territory of the USSR this front occupied a quasi-stationary position. On the map for 0700, 28 September (figure 6) a wave disturbance could have been detected on this front, over southern Byelorussia. Let us note, by the way, that a secondary cold front was noted to the northwest of the indicated frontal zone, which, as in the first case, undoubtedly played a part in the formation of the wave. The question naturally arose, will this wave develop and what will be its effect on the strengthening of the wind? The map of the relative topography 500/1000 millibars for the morning of 25 September showed that the temperature contrast in the high frontal zone was relatively small, in all only 16-20 dekabars per 1000 kilometers; that is, less than critical value for the main frontal zone. The wind velocities, even if near critical value, were considerably weaker than in the case of 23 September--that is to say, the wind velocity was equal to 50 kilometers per hour at a height corresponding to 70 millibars, and about 80 kilometers per hour at a height corresponding to 500 millibars, as contrasted with the corresponding 80 and 100 kilometers per hour of 23 September.

In altitudes up to 1 kilometer the wind reached 30 kilometers per hour. Thus, the difference between the absolute values of the wind velocity and the measurements of velocity as it varies with altitude has shown that these two cases of wave formation on arctic fronts must also differ in their effectiveness. In the case of 23 September we had to do with a very active frontal zone, and its activity was especially significant in the lower layer up to 3 kilometers, in which a discontinuity in wind velocity from 20 kilometers per hour at the earth's surface to 80 kilometers per hour at a height corresponding to 700 millibars was observed. In the case of 28 September the frontal zone was in general much less active than in the case of 23-24 September particularly in the lower layers, where the discontinuity of wind velocity in the layer from 1 to 3 kilometers was in all only about 20 kilometers per hour. From the other direction, this frontal zone evidently had a large store of energy at a height from 3 to 5 kilometers, since in this layer the discontinuity of wind velocity was about 30 kilometers per hour. The wave on the arctic front of 28 September did not develop to this extent since the air was not sufficiently cold at its rear; it remained a stable wave, quickly shifting along the frontal zone to the northeast. The speed of displacement of the wave was about 50-60 kilometers per hour; that is, it was about equal to the wind velocity at a height corresponding to 700 millibars. The direction of the displacement of the center of the wave, as in the first instance, almost exactly coincided with the position of the warm sector of the front. Regardless of the weak activity of the frontal zone, as well as the weak development of the wave, its effect on the deterioration of the weather was very real. The advance of this wave across the central regions of the European territory of the USSR did not noticeably strengthen the wind, but it caused abundant precipitation here. The precipitation fell chiefly ahead of the warm front, in front of which, judging by the above map, was located an area of current divergence. Very probably, therefore, the activity of the frontal zone in the way of heavy precipitation occurred in connection with that discontinuity of wind velocity which was observed in the layer of 3 to 5 kilometers.

Let us sum up our work in the following main conclusions:

1. Strong winds over the central areas of the European territories of the USSR on 23-24 September 1943 were caused by the advance of a young cyclone, formed on the evening of 22 September at the arctic front over the territory of Poland.

2. Very strong winds amounting to a hurricane were observed at the rear of the cyclone, when the latter reached the stage of a young (ideal) cyclone; the area of strong winds coincided therefore with the region of greatest positive tendencies beyond the cold front.

3. The main arctic front was very weakly manifested at the earth's surface; at the upper levels the corresponding frontal zone showed a temperature contrast of 20-25 dekabars per 1000 kilometers, wind velocities of 80-100 kilometers per hour (at a height of 3-5 kilometers), and a considerable discontinuity of wind velocity from the earth's surface up to 2-3 kilometers.

4. The formation of a wave on a quasi-stationary part of the arctic front was connected with the presence of a secondary cold front advancing somewhat north of the main front. Therefore, the wave thus formed was located near the delta of the high altitude frontal zone.

5. The center of the cyclone shifted in the direction of the isallobaric gradient, and the area of the greatest decrease of pressure was located near the center of the cyclone in the warm sector, ahead of the cold front, and the area of greatest increase was behind the cold front. The areas of decrease and increase of pressure were connected correspondingly with the zones of divergence and convergence of the upper currents.

6. The path of displacement of the center of the cyclone corresponded to the general position of the arctic front; in the course of the last 24-hour period the center of the cyclone was displaced in the direction of the warm sector of the arctic front, tending to bend from the south around the mass of arctic air; the length of the course of the cyclone for the 24-hour period was roughly equal to the length of the warm front.

7. The average speed of the cyclone from the initial stage of the wave to the stage of the young cyclone was about 60 kilometers per hour, which was 0.75 of the average speed of the main current at a height of 3 kilometers and 0.6 of the average speed at a height of 5 kilometers. Roughly the same relationship was maintained in the final stage of the cyclone's development.

8. In the analogous case of wave formation on the arctic front of 28 September 1943 the wave did not develop into a cyclone, as a result of the insignificant contrast of temperatures in the frontal zone, especially in the lower layers of the troposphere. The copious precipitation caused by the advance of this wave was evidently connected with the active anabatic gliding movement along the warm sector of the front in the layer from 3 to 5 kilometers, where a more significant discontinuity of wind velocity was observed.

9. Weakly manifested fronts on the synoptic map have a greater dynamic significance if they coincide with a sufficiently active high-altitude frontal zone, and conversely, their dynamic significance is not great if the corresponding high altitude frontal zone is not sufficiently active. But in both this and the other case the weakly manifested fronts have a great significance in the way of deterioration of weather in the lower levels of the troposphere.

10. In the analysis of the high-altitude maps it is still more necessary than in the analysis of surface maps to devote attention to the comparison of the data between separated stations, and also to the changes in data of stations from one date to another.

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TN: Map Symbols

В = high
Н = low
мПВ = maritime, polar, high
мАВ = maritime, arctic, high
ПВ = transitional
кПВ = continental, polar, high
ПрАВ = transitional, arctic, high

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IV. DETERMINATION OF CLOUD HEIGHTS BY HUMIDITY OF AIR AT GROUND LEVEL

by

A. N. Ippolitov

The height of clouds in the lower atmosphere has a direct influence on the operation of aviation. The presence of a solid overcast of very low clouds can bring about a complete suspension of the work of air transport. Knowledge of cloud heights, as well as the level of possible cloud formation, is very important for this reason. Unfortunately, determination of cloud heights by direct methods (pilot balloons and plane flights) has not yet become widespread. In most cases, cloud heights are determined visually, which inevitably leads to gross error in many cases. All this calls for investigation of new, even indirect, methods of determining cloud heights.

For the processes of adiabatic rise of air, taking place by virtue of vertical convection of a dynamic and thermal nature, the level of condensation of water vapors was calculated theoretically by Ferrel, who obtained the formula:

$$H = 122\ (t - \tau),$$

where t is the temperature of the air and τ is the dew point.

Empirical formula and diagrams of the dependence of cloud heights upon the humidity at ground level were obtained by direct observations of cloud heights (by pilot ballons). We refer here, for example, to the diagram of P. A. Molchanov, expressing the dependence of cloud heights upon the diversity of air temperatures and the dew point (P. A. Molchanov, Aerology, page 259, 1938).

From data of pilot-balloon observations of the Sverdlovsk observatory, we obtained a formula for the calculation of heights of convection stratus, fracto stratus, cumulus, fracto cumulus, and fracto nimbus clouds by the magnitude of relative humidity of air at ground level:

$$H = 22\ (100 - r),$$

where r is the relative humidity of the air (the journal, Weather, No 12, 1940). Comparison of cloud heights computed by Ferrel's formula and our's gives a sufficiently good correspondence, at any rate, up to a height of 1,000 meters. If we consider simplicity, convenience, and the small expenditure of time in calculating cloud heights by the formula $H = 22\ (100 - r)$, undoubtedly the latter is preferable, and the more so because it was obtained from factual material and, in great part, reflects the actual conditions in the atmosphere. Finally, it must also be pointed out that our formula is an approximate and not a functional one.

We have already indicated that Ferrel's formula and our formula were both calculated to compute cloud heights (as well as levels of condensation) forming in the rising air. Similar processes in the atmosphere are completed when there is a sufficiently developed vertical convection. Accordingly, it is also necessary to make these calculations when there is such an interchange (intermixture) of air that water vapors shift from the ground stratum to the level of their condensations (ie, up to the cloud level).

The absence of any direct instrumental determinations of the nature of the vertical convection is a fundamental and, for the time being, inevitable impediment to a broad application of methods of calculation of convection cloud heights. In many cases, however, we can judge the presence of a vertical convection by indirect indications.

For stratus, fracto stratus, and fracto nimbus clouds in the formation of which the dynamic convection (turbulence) of the air plays the chief role, wind with a velocity of not less than 3 meters per second can indicate a connection (interchange) of the lower levels of the air with the higher. For cumulus and fracto cumulus clouds, where a thermal convection is fundamental to their formation (a convection supported by the turbulence of the air at ground level), a daytime wind of not less than 3 meters per second will be the main indication of an interchange of air. However, these fundamental indications of convections are often unsatisfactory, and long with the observations, in each instance, it is still necessary to consider the nature of the weather at the time.

The altitudes of cumulus and fracto cumulus are the best indications of all, from the beginning of their formation in the morning until the time of their greatest development during the day (from 1400 to 1500). Good results for cumulus are obtained during the day, when these clouds have a normal daily variation; in the morning, separate clouds begin to appear in a clear sky, toward noon their number increases, and in the evening, they disappear or spread out in strato-cumulus vesperalis. With a clear sky at night, or early in the morning, the formation of dew or frost is generally observed. Their presence indicates absence of an interchange of air, and to determine the cloud heights (if there are any) under such circumstances is of no significance, for they will be erroneous, generally much lower than they actually are. Later, when the interchange will increase (the wind velocity increases and the dew vanishes), various broken cumulus clouds will begin to appear, and at this point, it will still be possible to calculate their heights. The relative humidity of the air when dew is formed is generally great, and after sunrise, it begins to fall rather quickly, slowing down at the time the first cumulus appear, although it continues to decrease until 1300-1500. The altitude of cumulus at this time is related to the relative humidity of the air. Afternoon, when the humidity increases, the altitude of cumulus no longer depends upon the highest relative humidity at ground level. The cumulus clouds observed at this time (let us say, after 1500), and later with their spreading and the strato-cumulus vesperalis, remain approximately at the altitude at which they were observed at noon, and their height may be calculated by the minimal value of the relative humidity of the air observed by day (1300-1500).

It is difficult to determine the heights of cumulus clouds not forming a solid overcast by the pilot-balloon method, since the balloon can enter a gap between clouds and be enveloped by them. The probability of the balloon's entering a cloud in the case of ragged forms of clouds (cumulus, fracto cumulus, fracto stratus and fracto nimbus) depends upon how overcast the sky is; with cloudiness, let us say, of 5/5 cumulus or 10/5 alto-stratus, fracto nimbus, there is a 50-percent probability of the pilot balloon's entering a cloud cumulus, fracto nimbus. It is sometimes difficult to tell whether the balloon has entered a cloud, or has been enveloped by it, and this leads to errors in determining altitudes. If cumulus develops significantly by day and changes to cumulo-nimbus with falling precipitation, the altitudes still can not be calculated--the humidity of the air at ground level will be affected by the fallen precipitation and will not be characteristic of the layers of atmosphere.

If, after the passing of heavy cumulo-nimbus precipitation with high wind velocities at ground level (let us say 5-10 meters per second) under the cumulo-nimbus layer, quickly vanishing fracto nimbus clouds are formed, their altitudes will be related to the values of the relative humidity, and the corresponding altitude calculations can be made.

Stratus, fracto stratus, and fracto nimbus clouds are formed, basically, when there is considerable turbulence in the terrestrial stratum of the atmosphere, and accordingly, the altitudes may be computed at any hour; but there must be a wind with a velocity of over 3 meters per second (at night, even 4-5 meters per second). Absence of precipitation is, therefore, also necessary. Best results for these forms of clouds are achieved at night, if there is no marked decrease of wind velocity from day to night (say, more than 3-4 meters per second), if the solid overcast of lower clouds is maintained at night, and if frost or dew are not formed.

A notable fact about strato-cumulus clouds is that their altitude may be determined by the magnitude of the relative humidity only during the day, only when they are not high (about 1,000 meters), when they have some cumulus formation of their various elements, and when these elements are fairly large and even increase (thicken) in the course of a day, which indicates an obvious relationship with the humidity condition at ground level.

In the majority of cases of high strato-cumulus clouds which are unrelated to the humidity condition at ground level, calculations can not be made. In general, we do not recommend calculating altitudes of strato-cumulus in reference to this condition.

The temperature of the air, when the formula $H = 22\ (100 - r)$ is used, is significant in that humidity of the air is less accurately determined when it is low (below minus 5 degrees), and in addition, it is not always clear from what value of saturating humidity it comes. Furthermore, at low temperatures and especially in winter, layers of terrestrial inversion of temperature are formed almost constantly over the continental stations (Siberia for example). These are intercepting layers impeding the transmission of water vapors upward from below. Calculations of cloud heights can not be made in reference to this at a temperature of under minus 5 degrees in the absence of temperature sounding.

Application of calculated formula without taking account of all the obstacles indicated here may lead to great errors in the determination of cloud heights. A large number of such errors were discovered in the materials of local geophysical observers, who produced, by order of the Central Board of the Hydrometric Service attached to the Council of Ministers, works examining the suitability of our formula for various areas of the USSR. For an average of about 70 percent of the calculations, the investigation was made under conditions in which our formula is not applicable.

Selecting instances corresponding to the conditions posited above, we calculated the average values of the coefficients of A in the formula H = A (100 - r) for 11 points in the USSR, the results of which appear in the following table:

Points	Gen No. Observations in Principle Material	No. Observations Not Corresponding to Method Conditions	No. Observations Used for Coefficient Calculations	Coefficient of A	Comment
Murmansk	40	22	16	25	--
Yakutsk	30	21	9	25	--
Vladivostok	101	83	13	24	--
Mashtash (Baku)	162	07	49	28	--
Omsk	82	164	16	20	--
Tashkent-Askhabad	114	96	17	22	--
Irkutsk	59	3	52	29	--
Kuybyshev	133	92	30	26	--
Aktyubinsk	122	71	45	21	--
Karaganda	155	93	54	21	--
Alma-Ata	94	21	27	22	--
Tbilisi	29	29	--	Not Calculated.	Observations made with wind at 4 m/sec.

As is evident from the table, we obtained values of the coefficients of A within the limits of 20-29 (average for all points 24). No dependence of the magnitude of A upon the geographical conditions is apparent, which actually, is as it should be.

In conclusion, let us point out the possibilities of the application of our formula.

It is quite possible to calculate the heights of low clouds in every single instance, but only with a competent approach to the matter. The use of the formula in question is particularly expedient for AMSG and the divisions of the Weather Service of the Local Boards of the Hydrometric Service, where a special map of low cloud heights may be made with the data of the network observations (even if they are 13-hour observations).

The formula may be used in climatology to calculate average cloud heights, their yearly path, and especially, the characteristic cloudiness of the little-investigated areas of airplane traffic. It is therefore necessary to bear in mind that we must not do detailed work on average values of cloudiness for any point, but must select instances of the presence of low clouds and take values of cloudiness appropriate to this period of observations.

The best may be the 13-hour period.

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V. VARIATIONS OF THE MONTHLY TOTALS OF THE HEAT OF SOLAR RADIATION

by

N. N. Kalitin

In order to study the changes in the inflow of the heat of solar radiation for any place from month to month and from year to year, there must be a considerable number of years of recording solar radiation without any gaps in the work. There is only one place on earth in which such a study is made possible by a series of observations on solar radiation over a considerable number of years. This is the Magnetic and Meteorological Observatory and Institute of Actinometry and Atmospheric Optics in Pavlovsk.

Recording of the intensity of solar radiation here was begun April 1912 and was continued without interruption until the end of August 1941, when the work was discontinued due to wartime conditions (direct approximation of the front to Pavlovsk). And so, for this point, there is a recording for 29 years, 5 months. It must be added that over this whole interval of time the recording was done with instruments of the same type (heliostat with a Savinov thermel and a point galvanograph), the entries were worked out by one and the same method, and they were supported by the standard pyrheliometer of Angström, which was repeatedly compared with the international standard. In this way, all the material is uniform and of full value. First-class, long-term series of Pavlovsk actinometric observations were repeatedly used for obtaining conclusions, not only by us (1, 2 and 5), but also abroad (Gorchinskiy and Kimball), and regularly published in the Bulletin of the Standard Actinometric Commission.

Before passing to an analysis of the variability of the monthly totals for the whole period of observations, we must consider the average magnitudes over a period of many years, as well as the maximum and minimum.

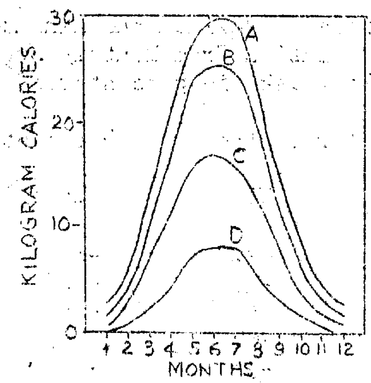
The conclusions on the variation of the monthly totals of the heat of solar radiation presented here were obtained on the basis of entries of the actinograph from 1914 to 1940, that is, over a period of 27 years. 1913 was excluded from the conclusions, since, in that year, the transparency of the atmosphere was still very low, as a result of the eruption of the Katmay volcano in Alyaysk 6 June 1912 (4). In table 1 are given the average monthly totals, both for the perpendicular and for the horizontal surface for 27 years:

Table 1. Average Monthly Maximum and Minimum Totals of Heat Radiation for Pavlovsk (in gram calories)

Mo.	Perpendicular Surface			Horizontal Surface		
	Avg	Max	Min	Avg	Max	Min
Jan	1,290	2,920	380	180	310	50
Feb	2,670	5,620	400	620	1,340	80
Mar	7,080	13,170	3,690	2,560	4,650	1,520
Apr	9,400	13,820	4,600	4,400	6,290	1,990
May	14,300	20,950	9,600	7,620	10,200	5,060
Jun	14,480	19,670	7,390	8,140	10,950	4,000
Jul	14,570	22,440	9,250	8,130	12,380	5,200
Aug	10,750	16,770	5,730	5,440	8,760	2,880
Sep	6,790	9,850	3,970	2,790	4,190	1,630
Oct	3,280	5,200	1,040	950	1,500	300
Nov	1,070	2,540	280	170	420	47
Dec	730	2,100	36	70	180	4

It is of interest to compare the totals of the heat of solar radiation that were actually obtained with the theoretical ones. A few years before this, S. I. Savinov drew up tables of the totals of heat of solar radiation for the horizontal surface for the northern hemisphere in a vacuum, and N. G. Evfimov made totals for the same surface in an ideal atmosphere.

Making use of their data, I calculated these totals for Pavlovsk for each month.



The results of such a computation are presented in figure 1. Curve A indicates the yearly course of the monthly totals of heat of solar radiation for Pavlovsk on 1 square centimeter of horizontal surface (in kilogram calories) in a vacuum; curve B is the same course in an ideal atmosphere; and curve C is for an unclouded sky with actual transparency of atmosphere. This curve was computed as an average from 718 unclouded days of reception of heat of solar radiation, for which there are actinograms for Pavlovsk from 1913 to 1940 (3). Curve D indicates the yearly course of the monthly totals of heat of solar radiation which was observed under natural conditions, that is, with a calculation of cloudiness also (the fourth column of table 1). Thus, the area between curves A and B indicates how much heat of solar radiation does not reach the earth's surface, being lost because of the ideal atmosphere; the area between curves B and C will give data on how weakened the reception of solar radiation is with a real, unclouded atmosphere (mainly on account of water vapor and dust); the area between curves C and D will show the weakening of radiation by cloudiness; and finally, the area bounded by curve D and the axis of the abscissae indicate the quantity of heat which reaches the earth's surface under actual conditions.

Figure 1 shows very graphically the role of the atmosphere in the weakening of the flow of solar radiation proceeding toward the earth's surface under Pavlovsk climatic conditions.

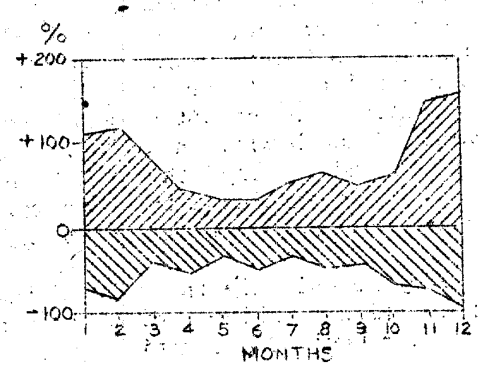
It is evident from the figures in figure 1 to what extent the monthly totals of the heat of solar radiation for one and the same month can vary for different years. These changes will be more graphically expressed in percentages.

Table 2. Deviations of Maximum and Minimum Totals of Heat of Solar Radiation From Monthly Averages for Pavlovsk (in percentages)

Month	Perpendicular Surface		Horizontal Surface	
	Max	Min	Max	Min
Jan	+126	-71	+106	-72
Feb	+111	-85	+116	-87
Mar	+ 86	-48	+ 82	-41
Apr	+ 47	-51	+ 43	-55
May	+ 46	-33	+ 34	-34
Jun	+ 36	-49	+ 35	-51
Jul	+ 54	-36	+ 52	-36
Aug	+ 56	-46	+ 61	-47
Sep	+ 45	-42	+ 50	-42
Oct	+ 59	-68	+ 58	-68
Nov	+137	-74	+147	-72
Dec	+238	-95	+157	-94

This was done in table 2, in which are shown the maximum and minimum deviations from the average of the monthly totals for both the perpendicular and the horizontal surface for the whole period of observation.

Table 2 indicates that the maximum and minimum magnitudes, differing slightly, have a clearly expressed yearly course for the perpendicular and horizontal surface. The greatest deviations of the monthly totals from the averages on either side approach the cold time of year, while the smallest deviations approach the warm time of year.



This yearly course is particularly evident in figure 2, in which the deviation for the horizontal surface is shown (in percentages). From year to year, considerably less variations of solar radiation are obtained in summer and spring than in winter for the climatic conditions of Pavlovsk (Leningrad); but for the winter months, the variations may be considerable.

The magnitudes they can reach are indicated by table 3, in which are presented the monthly totals of heat of solar radiation for each year for 27 years on a perpendicular surface for the same steady month (May) and the same variable month (December).

Table 3. Monthly Totals of Heat of Solar Radiation for May and December on a Perpendicular Surface (in Kilogram Calories)

Year	May	Dec	Year	May	Dec	Year	May	Dec
1914	12.8	0.4	1923	12.0	0.2	1932	12.8	0.6
1915	16.0	0.9	1924	12.1	0.9	1933	13.5	1.1
1916	11.2	0.4	1925	14.2	0.9	1934	12.4	0.7
1917	13.7	1.0	1926	11.2	1.1	1935	13.4	0.04
1918	15.0	0.2	1927	9.6	1.0	1936	14.7	0.5
1919	17.7	0.8	1928	11.9	0.7	1937	17.2	0.4
1920	17.9	0.6	1929	15.2	0.3	1938	10.9	2.1
1921	21.0	0.9	1930	15.3	0.8	1939	15.6	0.4
1922	11.2	0.8	1931	18.4	1.4	1940	18.2	0.4

The table indicates that the December totals, being considerably less in absolute magnitude than those of May, vary several times as much from year to year.

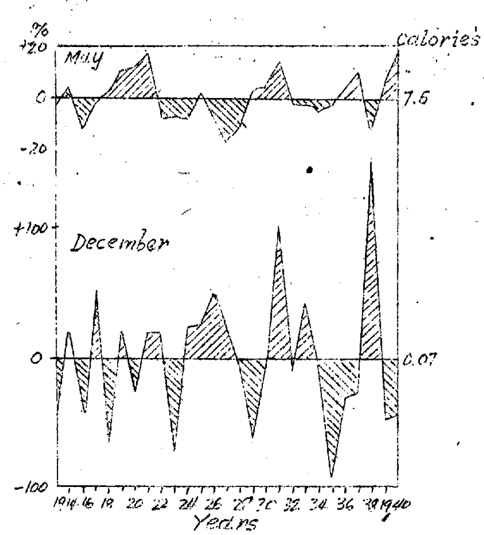


Figure 3 shows the deviation of the monthly totals from yearly averages for 27 years for a horizontal surface for May and December (in percentages). The figures in the drawing to the right indicate by median lines the average monthly totals on 1 square centimeter of horizontal surface (in kilogram calories)--- May, 7.6 calories, and December, 0.07 calories.

From the curves in the drawing, it is evident how much more the December totals vary than do the May totals.

Such an orderly, statistical treatment of the results of recording of solar radiation over a period of several years for places with different climatic conditions permits studying the problem of the steady values of solar radiation, which is of both practical and theoretical value.

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(TN: Asterisks indicate a Russian original.)

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VI. A NEW METHOD OF CALCULATING BALLISTIC WIND AND BALLISTIC TEMPERATURE

by

D. L. Laykhtman

In the year 1941, I submitted a new method of calculating ballistic wind and ballistic temperature. Since then, the method has been applied. In the article submitted here, the theory of the proposed method is explained. At the present time, ballistic wind is calculated for artillery purposes by the following formula:

$$\vec{V}_б = q_1\vec{V}_1 + q_2\vec{V}_2 + q_3\vec{V}_3 + \dots + q_n\vec{V}_n' \qquad (1)$$

where $q_i = \frac{t_i}{T}$ is the relative time that the projectile remains in the i^{th} layer; or using the average wind in the layer, by the formula:

$$\vec{V}_б = \vec{V}_{1\dots n-1} + \sqrt{n}(\vec{V}_{1,\dots n} - \vec{V}_{1\dots n-1}). \qquad (2)$$

The latter formula is less accurate, since, if the weight of all layers except the last one is calculated by it, they come out equal, and this is not actually the case. Another fault of formula (2) must be noted--mistakes in the ballistic wind caused by the indicated allowance differ from different wind velocities. However, the second method has been more widely applied because of its simplicity.

The method proposed below is unlike both indicated methods, it does not require any additional allowances (beyond the usual ones made in the solution of formula [1]), and it greatly simplifies the calculation of ballistic wind.

As is well known, in the solution of formula (1) and (2), all trajectories are divided into n layers of equal thickness; and therefore, when the ballistic wind is calculated, the speed of each layer with its weight contributes to the irregularity of the speed of the projectile. We shall divide the speed of the projectile into layers of equal thickness, and we shall thereby pick out the thickness of each layer in such a way that the ballistic coefficient (the weight of the layer) does not vary from layer to layer. It may be said beforehand that the thickness of the layer will diminish regularly with the altitude.

We shall show that, if the trajectory is broken up into n layers of equal weight, the thickness of each layer is found by the following formula:

$$\Delta_i = \frac{2(n - i) + 1}{n^2} Y, \qquad (3)$$

where i is the number of the layer (taking the ground layer as 1), and Y is the altitude of the trajectory.

Let $\frac{\Delta i}{Y} = a_i$, where $0 < a_i < 1$. Evidently a_i occurs together with i.

It is easy to show then that:

$$\vec{V}_6 = \sum_1^n \frac{t_i}{T} \vec{V}_i = (1 - \sqrt{1 - a_1})\, \vec{V}_1 + \\ + (\sqrt{1-a_1} - \sqrt{1-a_1-a_2})\, \vec{V}_2 + \dots + \\ + \sqrt{1-a_1-a_2-a_3 \dots -a_{n-1}}\, \vec{V}_n. \quad (4)$$

Actually, inasmuch as the length of the flight of the projectile on the trajectory of height Y is equal to $\sqrt{\frac{8}{g}}\sqrt{y}$, the relative time of the flight in the first layer is:

$$\frac{t_1}{T} = \frac{\sqrt{Y} - \sqrt{Y - \Delta_1}}{\sqrt{Y}} = 1 - \sqrt{1 - a_1}$$

In the second, it is:

$$\frac{t_2}{T} = \frac{\sqrt{Y - \Delta_1} - \sqrt{Y - \Delta_1 - \Delta_2}}{\sqrt{Y}} =$$

$$\sqrt{1 - a_1} - \sqrt{1 - a_1 - a_2}$$

And so forth.

Consequently, in order for the whole layer to have uniform weight, it must be:

$$1 - \sqrt{1 - a_1} = \sqrt{1 - a_1} - \sqrt{1 - a_1 - a_2} = \\ = \sqrt{1 - a_1 - a_2} - \sqrt{1 - a_1 - a_2 - a_3} = \dots = \frac{1}{n}.$$

By direct substitution, it is easy to prove that this equation is satisfactory, if $a_i = \frac{2(n - i) + 1}{n^2}$, by which equation (3) was also demonstrated.

We present a table of the values of a_i for different trajectories:

Serial No of Layer (i)	General Number of Layers (n) 1	2	3	4	5	6
VI	–	–	–	–	–	$\frac{1}{36}$
V	–	–	–	–	$\frac{1}{25}$	$\frac{3}{36}$
IV	–	–	–	$\frac{1}{16}$	$\frac{3}{25}$	$\frac{5}{36}$
III	–	–	$\frac{1}{9}$	$\frac{3}{16}$	$\frac{5}{25}$	$\frac{7}{36}$
II	–	$\frac{1}{4}$	$\frac{3}{9}$	$\frac{5}{16}$	$\frac{7}{25}$	$\frac{9}{36}$
I	1	$\frac{3}{4}$	$\frac{5}{9}$	$\frac{7}{16}$	$\frac{9}{25}$	$\frac{11}{36}$
Total	1	1	1	1	1	1

The ballistic wind is calculated very simply when the trajectory is divided into two layers.

We note in this case that, substituting a_1 and a_2 for the two layered trajectory in formula (4), we obtain:

$$\vec{V}_6 = \frac{\vec{V}_{3/4} + \vec{V}_{1/4}}{2}. \qquad (5)$$

Besides this, it is easy to show that if V is the average wind velocity in the whole trajectory, $\vec{V}_{1/4} = 4\vec{V} - 3\vec{V}_{3/4}$, then

$$\vec{V}_6 = 2\vec{V} - \vec{V}_{3/4}; \qquad (6)$$

This calculation is readily performed on Molchanov's circle, using a drawing scale applied to the main radius. The whole calculation is reduced to the following operations:

1. At the vertical angle corresponding to the apex of the trajectory, we find $\bar{V}$ on the drawing scale, bringing the horizontal angle up to the main radius, and we locate point 1, having computed $2\bar{V}$.
2. We bring the horizontal angle corresponding to three-fourths of the height of the trajectory up to the main radius and, opposite the corresponding angle on the measuring ruler, we locate point 2.
3. We calculate the ballistic wind by the vector connecting points 2 and 1. In this way, we avoid the necessity of using a supplementary instrument, Mikhaylovskiy's construction ruler. The ballistic wind is determined at once by the angles, without preliminary calculation of the true wind; and besides this, no new allowances are made (except

the usual ones made in the solutions of formula (1).

Calculation of ballistic wind by the proposed method, being more accurate in principle than the method using average speeds, is being adopted at the present time by the artillery service and reduces their work by about 50 percent.

The proposed method is also preferable for calculating temperature. As is well known, the ballistic deviation of temperature is calculated, in cases of layers of equal thickness, by the following formula:

$$\Delta\tau_{b_n}^{\circ} = \Delta\tau_1^{\circ}\frac{t_1}{T} + \Delta\tau_2^{\circ}\frac{t_2}{T} + \Delta\tau_3^{\circ}\frac{t_3}{T} + \ldots + \Delta\tau_n^{\circ}\frac{t_n}{T},$$

that is, the ballistic deviation must be determined in each layer, each of them must be multiplied by the corresponding relative time, and then an algebraic addition must be made.

The multiplying operation is dropped in the proposed method; it is necessary only to determine the ballistic deviation at the heights previously determined (at the centers of the layers), and their mean arithmetic addition must be found. The whole operation could be carried out in the mind, and the necessity of using Naumov's plane table is obviated.

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VII. A NET FOR FILLING RADIOSONDE BALLOONS UNDER TRAVELLING CONDITIONS

by

A. S. Stepanov

Filling radiosonde balloons under travelling conditions is very difficult, especially in windy weather, when there is no enclosed space or tent available.

The construction of a net is described here, facilitating the process of filling radiosonde balloons in the open air.

To make the net, it is necessary to connect 2, 3 or 4 cords 480-500 centimeters long, about equal to the length of the circumference of the balloon. The cords are all tied together in the middle, and a separate cord is tied to the knot thus obtained. At the free ends of the cords, loops are made. The middle parts of the free ends are fastened by the separate cord, and 120 centimeters are measured off on the cord, as indicated in figure I, if 2 cords are connected for making the net. If it is 3 cords, these intervals must be 80 centimeters, and if 4 cords, 60 centimeters. The free cords release the ends of the transverse one, and the ends are connected before the balloon is filled. Before releasing it, they are untied again.

Before the balloon is filled, the net is hung on 4 pegs about 2 meters long. On top of two longer pegs, the crossbeam is placed.

The top of the net is drawn up to the top with a cord thrown over the crossbeam. The way the net is placed on the filled balloon is shown in figure II. The hanging, free ends of the cords' loop are connected by a separate cord.

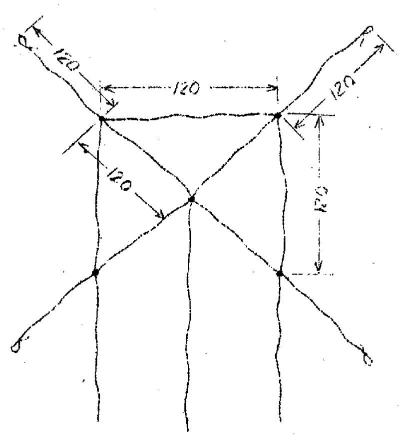

Figure 1

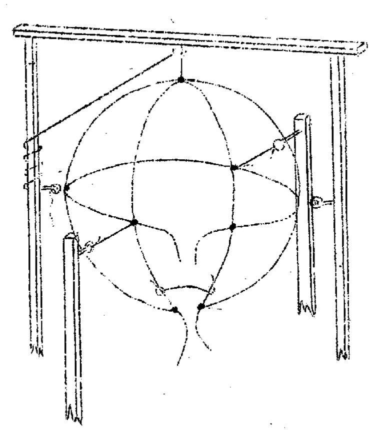

Figure 2

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